



Review Article

A tribute to Michael R. Raupach for contributions to aeolian fluid dynamics



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ABSTRACT

Since the pioneering work of Bagnold in the 1940s, aeolian research has grown to become an integral part of earth-system science. Many individuals have contributed to this development, and Dr. Michael R. Raupach (1950–2015) has played a pivotal role. Raupach worked intensively on wind erosion problems for about a decade (1985–1995), during which time he applied his deep knowledge of turbulence to aeolian research problems and made profound contributions with far-reaching impact. The beauty of Raupach's work lies in his clear conceptual thinking and his ability to reduce complex problems to their bare essentials. The results of his work are fundamentally important and have many practical applications. In this review we reflect on Raupach's contribution to a number of important aspects of aeolian research, summarise developments since his inspirational work and place Raupach's efforts in the context of aeolian science. We also demonstrate how Raupach's work provided a foundation for new developments in aeolian research. In this tribute, we concentrate on five areas of research: (1) drag partition theory; (2) saltation roughness length; (3) saltation bombardment; (4) threshold friction velocity and (5) the carbon cycle.

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1. Introduction

Aeolian research is multi-disciplinary, but its core lies arguably in the fluid dynamic interactions between soil particles, the atmosphere, and the soil surface. Since the early work of [Bagnold \(1941\)](#), it has advanced to become an integral part of earth-system studies. Aeolian processes are highly relevant topics in the earth sciences because of the need to: (1) better quantify the dust cycle for climate projections; (2) assess the anthropogenic impact on natural and human environments; (3) prevent soil loss from wind erosion in land-conservation practice; and (4) understand aeolian processes and landform development on other planets in particular, Mars and Venus, as well as moons such as Titan. Many individuals have contributed to this development, and Dr. Michael R. Raupach (1950–2015) was one of the most outstanding ([Steffen, 2015](#)).

For colleagues in aeolian research, and in climate research at large, Michael R. Raupach is Mike, but he used to abbreviate his name MR². This abbreviation was related to his university training in Applied Mathematics. Raupach received his BSc degree, with honours in mathematical physics, from the University of Adelaide in 1971, and a PhD in micrometeorology (under the supervision of Prof. Peter Schwerdtfeger) from the Flinders University of South Australia in 1976. After a postdoctoral position at the University of Edinburgh, he joined the Centre for Environmental Mechanics (CEM, also referred to as the Pye Lab) of the CSIRO (Commonwealth Scientific and Industrial Research Organisation) in Canberra in 1979, where he worked for much of his 35-year career. From 2000 to 2008, he was inaugural co-chair of the Global Carbon Project, an international program bridging the research effort between the natural and human dimensions of the carbon cycle. In February 2014, he took up the role of Director at the Climate Change Institute of the Australian National University and remained an Honorary Fellow with the CSIRO. Based on his research foci, his career can be divided into two stages. In the first he worked on atmospheric boundary-layer turbulence and atmosphere-land-surface exchanges, including aeolian processes, and in the second on climate change, in particular the carbon cycle.

Raupach's scientific drive originated from his passion for protecting the environment, and his interest in aeolian processes following from his concerns with land conservation. The period of 1977–1988 saw three successive El Niño events, including the intense phase of 1982–1983, which brought record drought to eastern Australia, turning the farmlands in the wheat-sheep belt into a hot spot of wind erosion. On 8 February 1983, a “cool change” (a dry cold front) preceded by hot (43.2 °C) gusty north-

erly winds blew large quantities of red-brown dust over Melbourne. This event inspired Raupach to write one of his first essays on wind erosion ([Raupach et al., 1994](#)), which was pioneering in its attention to three fundamental goals of dust research: identification of dust sources; estimating dust loads; and quantifying the nutrient loss of topsoil by wind erosion. Their estimate of the dust loading (2 ± 1 Mt) in the 1983 Melbourne dust storm was one of the earliest attempts to quantify event-based dust loading. This value was based upon a few back-of-the-envelope calculations; reducing a complex problem to its fundamental components, for which Raupach became famous. Raupach's estimate of topsoil nutrient loss was highly innovative, and 20 years later, wind-erosion related soil nutrient and soil carbon transport has become one of the most fundamental aspects of studies on the dust cycle.

In 1985, John Leys, then with the New South Wales Soil Conservation Service, had just started his PhD at Griffith University in Brisbane under the supervision of Professor Grant McTainsh, and was developing a portable wind tunnel for wind-erosion field experiments. At the time, Raupach was among a group of world-class micro-meteorologists gathered in the Pye Lab, conducting wind-tunnel experiments on flow over complex terrains. Raupach and Leys went on to modify the design of [Marsh and Carter \(1983\)](#) and develop Australia's aeolian-research wind tunnel ([Leys and Raupach, 1990](#)). The excellent fluid dynamic features of this tunnel made it a valuable research tool not only for land-conservation studies ([McTainsh and Leys, 1993](#)), but also for the studies of basic wind-erosion processes ([Shao and Raupach, 1992](#); [Shao et al., 1993](#)). In 1991, a group of Australian wind-erosion researchers gathered at the Murdoch University in Perth and staged the 1st Australian workshop on wind erosion ([Fig. 1](#)). In this workshop, William Nickling gave a keynote presentation “Shear Stress: What Drives Wind Erosion Processes”. Following the meeting, with a cool sea breeze and bright stars in the sky in the port of Fremantle, Raupach treated everyone with beer. In 1993, the group met again in the Mallee country town of Mildura and formed the Wind Erosion Research Community of Australia (WERCA, a name that Raupach and Grant McTainsh conceived over drinks at the meeting). Dale Gillette gave a philosophical talk on the paradigms of wind erosion. It is unfortunate that Raupach will not be with us for the ninth International Conference on Aeolian Research (ICAR IX) to be held in Mildura in 2016. However, the influences of his work will be evident at the conference and will provide a legacy for a considerable time.

Raupach worked for about a decade (1985–1995) intensively on wind erosion problems, but he did so brilliantly by relating aeolian



Fig. 1. Michael R. Raupach (back, 6th left) among the participants of the 1st Australian Workshop on Wind Erosion, 1991, Murdoch University, Perth. Several contributors to this paper were among the participants: Grant McTainsh (front, 1st left), Paul Findlater (front, 2nd left), Yaping Shao (back, 1st left), William Nickling (back, 5th left), John Leys (back, 7th left). The workshop convener was William Scott (front, 3rd left).

problems to his deep knowledge of turbulence, and made profound contributions to the field with far-reaching influences and a lasting legacy. The beauty of Raupach's work is crystal clear conceptual thinking, reducing problems to their essentials and expressing that essence with elegance yet simplicity. The results of his work are robust and practically applicable. In this review we reflect on Raupach's contribution to a number of important aeolian research themes, summarise the developments since his inspirational work and place Raupach's effort in the context of aeolian science. We also demonstrate how Raupach's work provided many foundations or platforms for the development of his work and the investigation of new research. For brevity, we will concentrate on Raupach's work in five areas: (1) drag partition theory; (2) saltation roughness length; (3) saltation bombardment; (4) threshold friction velocity; and (5) carbon cycle.

2. Drag partition theory and applications to wind erosion studies

2.1. The Raupach drag partition theory

In the atmospheric surface layer, the profile of the mean wind is approximately logarithmic in form and the shear stress, τ , also referred to as drag, is vertically approximately constant. Thus, the flow in the surface layer is characterised by the aerodynamic roughness length, z_0 , and the shear stress that is often expressed in terms of friction velocity, $u_* = \sqrt{\tau/\rho}$, with ρ being air density.

At the second International Conference on Aeolian Research (ICAR II) in Denmark (1990), Raupach, Gillette and Leys discussed the difficulties of sand flux modelling in the shrub lands of Texas and Australia as opposed to the sandy beaches of Denmark. They soon realised that in many wind-erosion applications, the knowledge of τ alone is insufficient, as in shrub lands the shear stress on the intervening erodible surface, which drives the sand movement, is subject to the influences of the shrubs. Raupach generalised this discussion to the fluid dynamic problem of drag partition over rough surfaces, i.e., the partition of the total drag

into a pressure drag on roughness elements and a friction drag on the surface. Raupach (1992) laid the foundation of the drag partition theory and Raupach et al. (1993) demonstrated how this theory can be applied to estimating sediment transport threshold over various rough aeolian surfaces. Raupach's work led the way to numerous studies that followed, ranging from wind-tunnel and field experiments, numerical modelling, remote sensing and theory. We know today that it is desirable in general to treat τ in wind erosion applications as a stochastic variable and to statistically quantify its spatial and temporal variations. As discussed later in this review, the spatial variability of shear stress is a critical part of heterogeneous aeolian processes, while its temporal variability is important for intermittent saltation and dust emission.

Shear stress variation in nature can be very complicated, and simplifications are necessary for theoretical analysis (Lettau, 1969; Arya, 1975). Following Schlichting (1936), Raupach (1992) suggested that a rough surface can be considered to be composed of roughness elements (in the spirit of Raupach's analysis, it seems appropriate to invent the word "roughness") superposed on a smooth substrate surface. The total drag is thus expressed as:

$$\tau = \tau_r + \tau_s \quad (1)$$

where τ_r is the drag on the roughness elements, or pressure drag, and τ_s the drag on the substrate surface, or surface drag. The task of drag partition is to determine the ratios τ_r/τ and τ_s/τ , and to estimate how these ratios depend on the roughness characteristics. An immediate question that arises is how the surface roughness can be quantified. Raupach aimed to find an analytical solution for the simplest case possible and thus assumed that the surface consists of randomly distributed cylinders uniform in size, each having a frontal area of a_f . It follows that if the number density of the roughness elements is n (number per unit area), then the frontal area index of the roughness elements is

$$\lambda = n \cdot a_f \quad (2)$$

which is the only input parameter for the Raupach (1992) scheme. This conceptual simplification was influenced by the work Raupach was very familiar with, in particular the wind-tunnel experiments of Marshall (1971) and Wooding et al. (1973), all from the Pye Lab.

Raupach (1992) introduced the concept of effective shelter area, A , and volume, V , associated with an individual roughness element (Fig. 2), and made two hypotheses:

Hypothesis I. For an isolated roughness element of breadth b and height h in a deep turbulent boundary layer with friction velocity u_* and mean velocity U_h at height h the effective shelter area A and volume V scale as:

$$A \sim bhU_h/u_* \tag{3a}$$

$$V \sim bh^2U_h/u_* \tag{3b}$$

Hypothesis II. When roughness elements are distributed uniformly or randomly across a surface, the combined effective shelter area or volume can be calculated by randomly superposing individual shelter areas or volumes.

With these hypotheses, Raupach (1992) found that

$$\frac{\tau_s}{\tau} = \frac{1}{1 + \beta\lambda} \tag{4a}$$

$$\frac{\tau_r}{\tau} = \frac{\beta}{1 + \beta\lambda} \tag{4b}$$

with $\beta = C_r/C_s$, where C_s is the frictional drag coefficient and C_r the pressure drag coefficient. Eq. (4) is a simple yet robust model supported by the wind-tunnel measurements of Marshall (1971) as well as the numerical simulations of Li and Shao (2003).

The results of Raupach (1992) have two immediate applications, first to estimate threshold friction velocity for wind erosion, u_{*t} , and second to estimate aerodynamic roughness length, z_0 . Suppose for a surface σ is the ratio of roughness basal area to frontal area, then the exposed fraction of the surface subject to wind erosion is $(1 - \sigma\lambda)$, and the shear stress on the exposed surface is:

$$\frac{\tau'_s}{\tau} = \frac{1}{(1 - \sigma\lambda)} \frac{1}{(1 + \beta\lambda)} \tag{5}$$

Here, τ'_s is the spatially averaged stress on the exposed surface. If we assume the largest stress acting on the surface is τ''_s and $\tau''_s(\lambda)$ equals $\tau'_s(\lambda_0)$ with $\lambda_0 = m\lambda$ and $m < 1$, then we have:

$$R_t^2 \equiv \frac{\tau''_s}{\tau} = \frac{1}{(1 - m\sigma\lambda)} \frac{1}{(1 + m\beta\lambda)} \tag{6}$$

R_t is the ratio of u_{*ts}/u_{*tr} , with u_{*ts} being the threshold friction velocity for the surface free of roughness and u_{*tr} for the rough surface. It follows that:

$$u_{*tr} = u_{*ts} \sqrt{(1 - m\sigma\lambda)(1 + m\beta\lambda)} \tag{7}$$

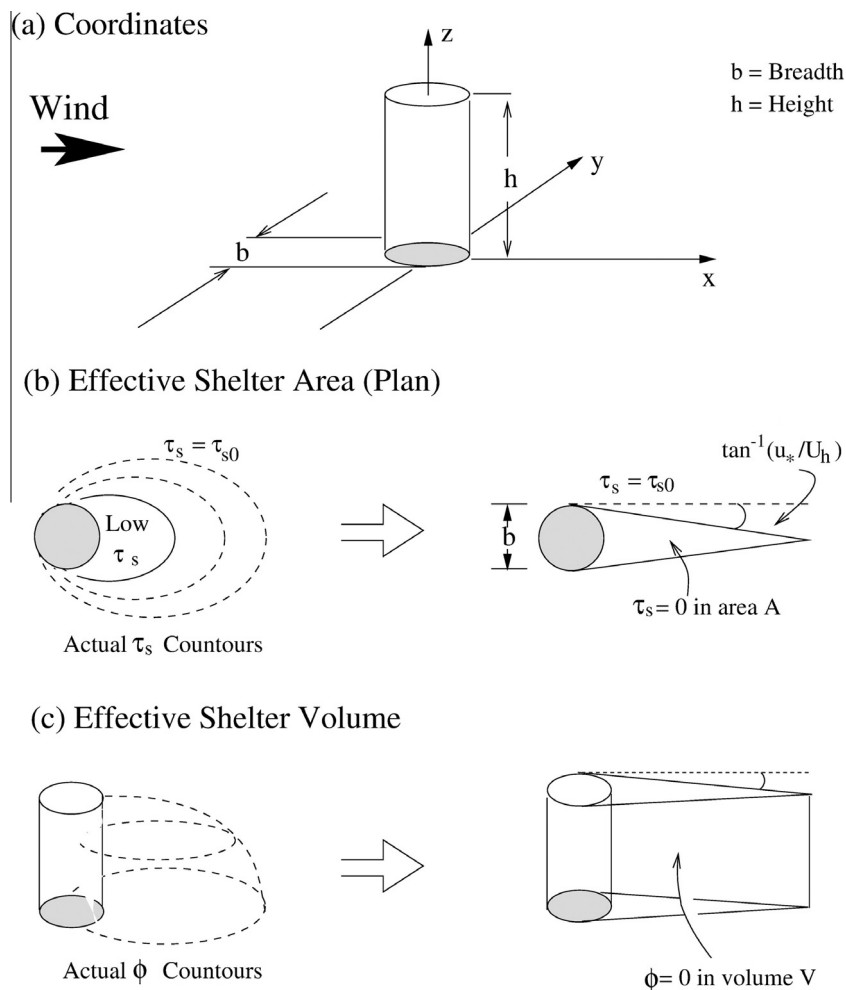


Fig. 2. Raupach's conceptual model for drag partitioning. A rough surface is considered to consist of roughness elements and a substrate surface. A roughness element produces an effective sheltering area and volume. The integrative effect of the roughness elements can be estimated by random superposition [Redrawn from Raupach (1992)].

Eq. (7) provides a simple way for the correction of u_{*t} for rough surfaces, and its validity is confirmed through comparison with the existing data of Gillette and Stockton (1989), Musick and Gillette (1990), Lyles and Allison (1976), Iversen et al. (1991), Crawley and Nickling (2003), Li and Shao (2003), and Sutton and McKenna-Neuman (2008).

Raupach (1992) and Raupach et al. (1993) provided for the first time a strong theoretical underpinning for explaining the impacts of roughness elements on aeolian thresholds and fluxes and a deep insight into the aeolian fluid dynamics. The method used in Raupach (1992) is unique in that by introducing the sheltering area and volume, Raupach took a “quantum fluid” approach, in that he discretely quantified the effect of an individual roughness element and then estimated the total effect of all roughness elements through random superposition. For this reason, it is appropriate to call roughness elements roughtons.

2.2. Wind-tunnel and field experiments on drag partition

To test the theory of Raupach (1992) and Raupach et al. (1993), William Nickling and Jack Gillies thought it critical to: (1) bridge theory to field measurements at the full scale; (2) examine how R_t behaves if roughness elements are real plants; and (3) investigate the impact of roughness elements on saltation transport, in addition to mean u_{*t} . They carried out field and wind-tunnel studies to evaluate C_r as a function of the Reynolds number, R_e , for different plants (Gillies et al., 2000, 2002). At the USDA Jornada Experimental Range (Gillies et al., 2006, 2007), they placed staggered arrays of large cylindrical roughness elements on a bare open surface and instruments between them to measure the total drag, surface drag, pressure drag and sand fluxes. It was found that C_r for plants is both plant-form and R_e dependent. This implies that drag partition for surfaces with plants is not necessarily fixed, but changes as the plants reconfigure themselves in response to wind. The more flexible the plant, the greater the proportion of shear stress acting on the substrate surface, and C_r declines with R_e . This finding implies that steppe landscapes (Shinoda et al., 2011), which

are typically composed of grass-type species, are likely more erodible than the shrub-dominated landscapes of the southwestern US deserts (Gillette and Pitchford, 2004; Gillette et al., 2006; King et al., 2005). It was also found that while sand flux scales with λ , it is also dependent on the height of the roughtons. For the same λ , elements with $h \geq 0.3$ m are more effective in reducing sand flux than shorter elements, e.g., $h \leq 0.1$ m (Gillies et al., 2006, 2015; Gillies and Lancaster, 2013). These experiments show that the Raupach et al. (1993) model performs well in general, but additional considerations should be given to roughness configuration to fully account for the observed saltation flux variations over rough surfaces. While Eq. (7) has three parameters, m , σ and β , it appears sufficient to choose appropriate β values (between 100 and 400) to fully describe the observed R_t for a wide range of surfaces, but to keep m and σ constant [e.g., 0.5 and 1, respectively, as set Raupach et al. (1993)], as Fig. 3 shows.

The simplicity of Eq. (7) is a strength for its application, in that it requires only a few measurable parameters (Wolfe and Nickling, 1996; Lancaster and Baas, 1998; King et al., 2006). This approach is widely used today in wind erosion models. In some studies e.g., Marticorena and Bergametti (1995), the drag partition scheme of Arya (1975) based on roughness length is used. However, because roughness length is closely related to roughness configuration (e.g., frontal area index, λ), the schemes of Arya (1975) and Raupach (1992) and Raupach et al. (1993) are in essence equivalent (see also, Raupach, 1994).

2.3. Extension of drag partition theory

Real aeolian surfaces are much more complex than is assumed in Raupach (1992) and Raupach et al. (1993). For practical applications, the Raupach (1992) theory requires several extensions: (1) the validity of Eq. (4) is limited to about $\lambda \leq 0.1$, but natural surfaces often have much larger roughness densities; (2) for surfaces with larger λ , it is not clear how shelter areas and volumes can be evaluated and how they superpose due to the interactions among the turbulent wakes associated with the roughness

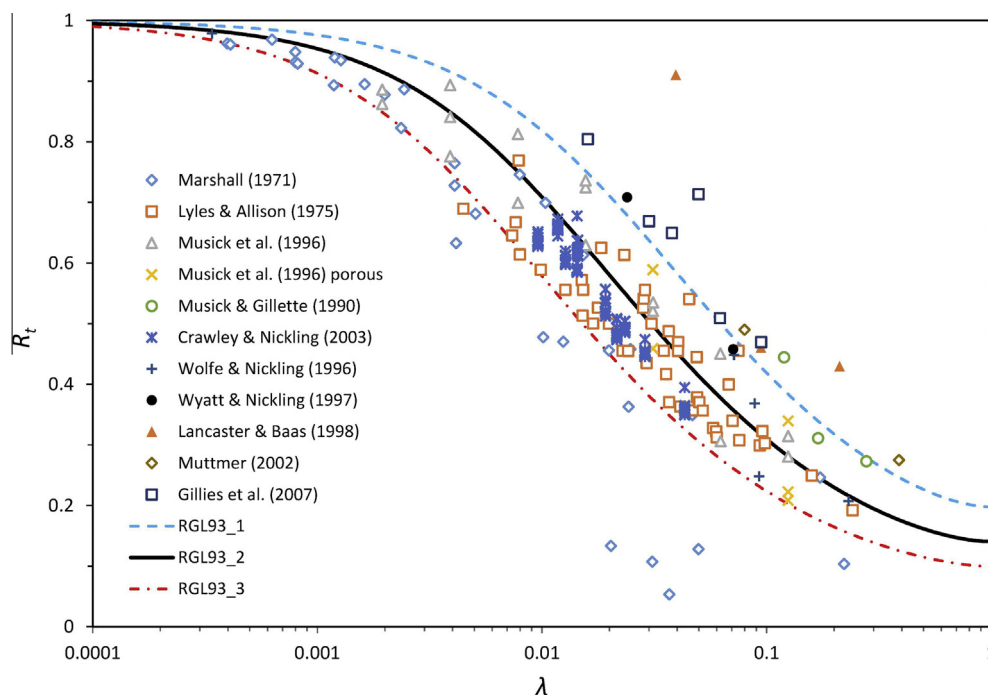


Fig. 3. A compilation of R_t versus λ data from wind-tunnel and field experiments (symbols). RGL93_1, RGL93_2 and RGL93_3 are the estimates using the Raupach et al. (1993) scheme with $m = 0.5$, $\sigma = 1$, and $\beta = 100, 200$ and 400 , respectively.

elements; (3) there are large uncertainties in the parameters β and m , as both are dependent on the roughness-element properties (e.g., porosity and elasticity) and configuration (arrangement and aspect ratio).

It is possible to derive the results of Raupach (1992) with simpler assumptions. For instance, it is sufficient to assume linear superposition of shelter areas and volumes instead of random superposition as applied in Raupach (1992) (Shao and Yang, 2008). More generally, we can write

$$\tau_s = \rho f_s C_s U_h^2 \quad (8a)$$

$$\tau_r = \rho f_r C_r \lambda U_h^2 \quad (8b)$$

where f_r and f_s are modification functions of the respective drag coefficients arising from the interacting flows shed by roughness elements. Eq. (8) leads to Eq. (4), subject only to $f_s = f_r$, which is $\exp(-c\lambda U_h/u_*)$ in Raupach (1992).

Eq. (1) is appropriate for small roughness density, but as λ increases, the total drag on the rough surface, τ , is better written as:

$$\tau = \tau_r + \tau_s + \tau_c \quad (9)$$

where τ_c is the friction drag on the surfaces of roughness elements. As $\lambda \rightarrow \infty$, we expect $\tau_r/\tau \rightarrow 0$ but Eq. (4) states that $\tau_r/\tau \rightarrow 1$ due to the neglect of τ_c . In general, the total drag can be partitioned into three components following Eq. (9) and the individual terms expressed as:

$$\tau_r = \rho f_r C_r \lambda U_h^2 (1 - \eta) \quad (10a)$$

$$\tau_s = \rho f_s C_s U_h^2 (1 - \eta) \quad (10b)$$

$$\tau_c = \rho C_s U_h^2 \eta \quad (10c)$$

where η is fraction of cover and f_r and f_s are functions of λ and η represent the modifications to C_r and C_s arising from the interactions of the turbulent wakes of roughness elements. With this formulation, the drag partition problem is now reduced to determine f_r and f_s . It is also found in Shao and Yang (2008) that

$$\frac{u_*^2}{U_h^2} = f_r \lambda (1 - \eta) C_r + [f_s (1 - \eta) + \eta] C_s \quad (11)$$

that shows that u_*^2/U_h^2 is a weighted average of the pressure and surface drag coefficients. In neutral atmospheric boundary layers, we have $U_h = \frac{u_*}{\kappa} \ln\left(\frac{h-d}{z_0}\right)$ and Eq. (11) can be written as:

$$\kappa^2 \ln^{-2}\left(\frac{h-d}{z_0}\right) = f_r \lambda (1 - \eta) C_r + [f_s (1 - \eta) + \eta] C_s \quad (12)$$

Eq. (12) shows that the roughness length, z_0 , can be determined in terms of drag coefficients for a given zero-displacement height, d . Thus, in a drag partition theory, we actually make two inter-related statements. The first is about the behaviour of drag partition functions; and the second about the behaviour of u_*/U_h or equivalently a statement on the drag coefficients or on the roughness length, z_0 . The above formulation of Shao and Yang (2008) as an extension of Raupach (1992) reduces the drag partition problem to the determination of the drag coefficients modification functions. The Shao and Yang scheme requires both frontal area index and fraction of cover as input parameters.

Another extension of Raupach (1992) was made by Okin (2008). The Okin scheme builds on the basic insight that τ_s in the lee of a roughness increases with distance downwind. While Raupach (1992) expressed the wake effect by means of shelter area and volume, the Okin scheme takes a probabilistic approach that envisions the surface to be made up of points that are some distance down-

wind of a roughness. The shear stress experienced at each point is an increasing function of this distance, scaled by the height of the roughness, multiplied by τ . With this approach, the frontal area index is no longer the best variable for characterising vegetation cover, but is replaced by the separation distance between the roughness elements. In Okin (2008), the shear stress on the soil surface is variable across the landscape, as originally envisioned in Raupach (1992). This approach allows some areas of the surface to experience transport while the more protected areas do not. This approach differs from that of Raupach et al. (1993) in which the threshold shear stress is seen to be a property of the bulk surface. As a result, the Okin scheme is able to predict transport even at relatively high vegetation cover, in accordance with field observations. Several studies published since have supported this approach (Webb et al., 2014; Walter et al., 2012a,b; Li et al., 2013).

2.4. Saltation heterogeneity

At ICAR-V (Lubbock, USA, 2002), Raupach incisively exposed the challenges of large scale aeolian modelling. Raupach and Lu (2004) subsequently published a review of land surface processes on aeolian transport modelling during the previous two decades and identified four challenges: (1) the fidelity of process representation; (2) upscaling point-scale process models in the presence of unresolved heterogeneity in space and time; (3) availability of spatial data for specifying model inputs and boundary conditions; and (4) large-scale parameter estimation. To date, these challenges remain largely unresolved but the explicit and clear articulations in Raupach and Lu (2004) provide essential guidance for what needs to be done.

Raupach and Lu (2004) suggested that improvements should be made in point-scale parameterisations including "...the effects of crusts and surface cohesion leading to supply-limited saltation and dust uplift, deposition to sparse vegetation...". There have been some developments in this area in particular with investigations of soil moisture (e.g., Wiggs et al., 2004), the use of laser scanning technology to describe small-scale roughness dynamics (e.g., Nield et al., 2013), angular reflectance measurement and bi-directional reflectance modelling to characterise changes in soil condition in space and time (Chappell et al., 2005, 2006, 2007) and retrieval of roughness changes in space and time using satellite remote sensing (Wu et al., 2009).

Raupach and Lu (2004) characterised point-scale transport models as, $f = f(v)$, where f is a flux and v is a vector of control variables. Part of the challenge with aeolian transport models is that they require flux and driving variables averaged in space and time. Raupach and Lu (2004) first defined

$$\bar{f}(\bar{v}) = \int f(v)p(v)dv \quad (13)$$

where $p(v)$ is the probability density function (PDF) of v . If $f(v)$ is linear, then $\bar{f}(\bar{v})$ has the same form as $f(v)$. Raupach and Lu (2004) considered the cases when $f(v)$ is highly nonlinear (e.g., involving threshold responses), which originates from the interaction between the nonlinearity in $f(v)$ and statistical variability in v that causes the upscaling problem to be mathematically nontrivial and dependent on sub-grid-scale variability through the PDF $p(v)$. Raupach and Lu (2004) used an example of heterogeneous vegetation to show profoundly that "...major errors arise from upscaling procedures, which neglect the interaction between model nonlinearity and statistical variability in driving variables". They also demonstrated that even a first approximation to the sub-grid-scale variability can lead to substantial improvement in flux estimates.

The effect of surface heterogeneity, i.e., the deliberately neglected “level of details” in Raupach et al. (1993) has been subject to intensive studies in more recent years, as it has been identified as a significant source of uncertainty in its application. Yang and Shao (2005) demonstrated that in case of very small roughness density, the shear stress variability due to the presence of roughness actually enhances rather than suppresses wind erosion. Raupach and Lu (2004) recognised that while no sediment trans-

port is predicted at large λ , this may happen in reality depending on roughness configuration (Okin, 2008), and that accounting for the PDF of λ can improve the model estimates, although this can be practically difficult (Walter et al., 2012a,b; Dupont et al., 2013, 2014).

Brown et al. (2008) conducted wind-tunnel experiments to determine the PDF of τ_s and R_t for a range of λ , roughness configurations, and free-stream wind velocities, U_i . The authors

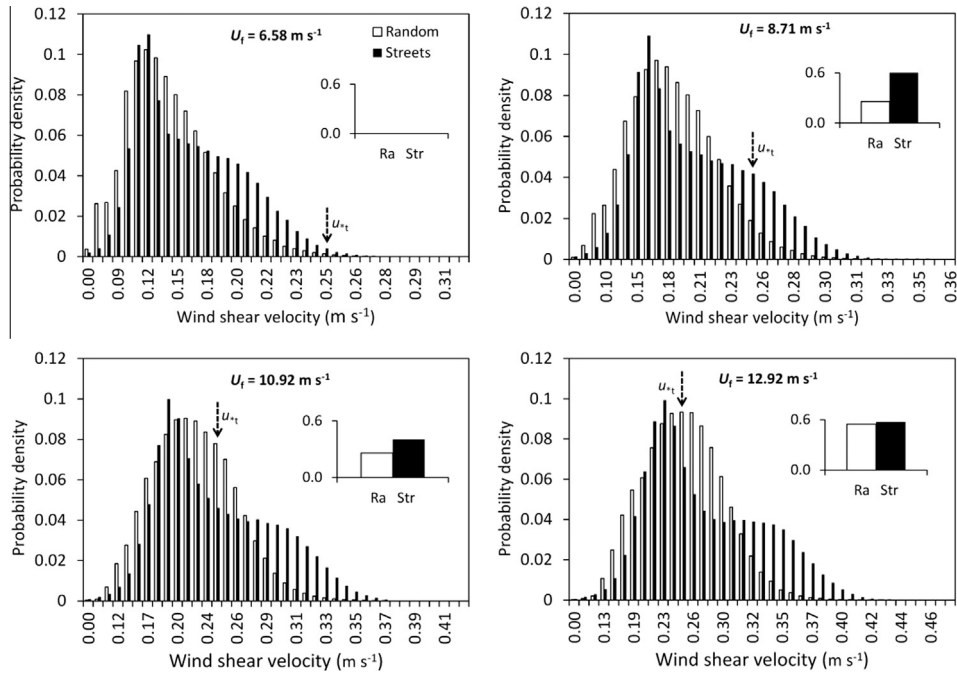


Fig. 4. Histograms illustrating the effect of the ‘random’ and ‘street’ roughness configurations on wind shear velocity (u_w) calculated from measured surface shear stress (τ_s) distributions at a roughness density $\lambda = 0.1$ and four free stream wind velocities (U_i). Inset graphs show the proportion of τ_s greater than a threshold shear velocity $u_{*t} = 0.25 \text{ m s}^{-1}$ for the random (Ra) and street (Str) configurations. These proportions are indicative of the relative sediment fluxes produced for the two roughness configurations.

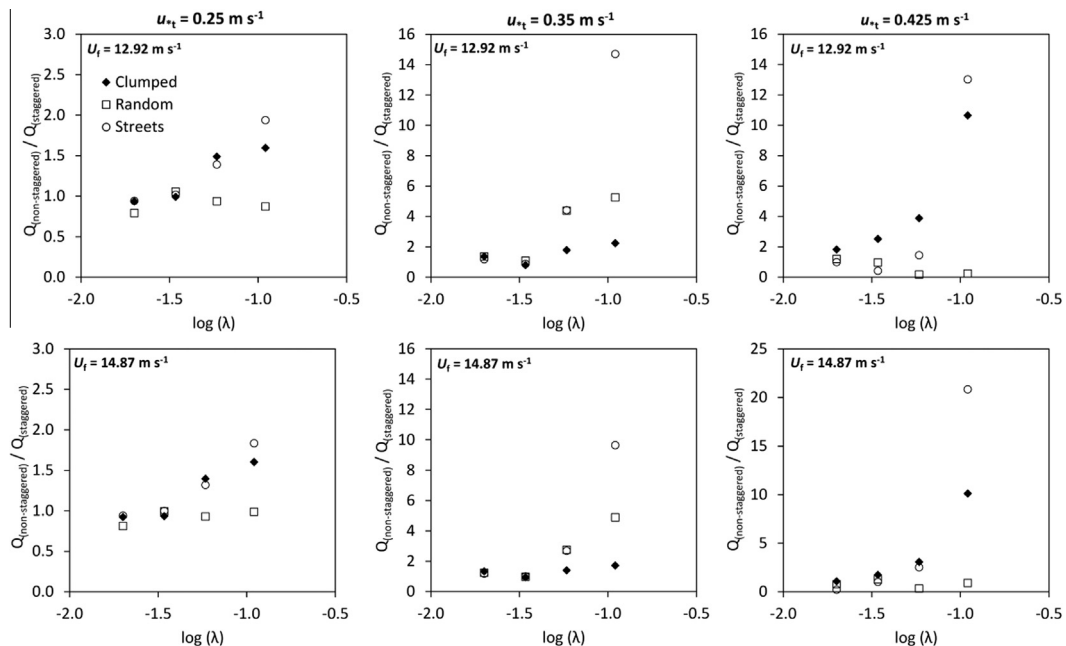


Fig. 5. Graphs showing roughness configuration effects on horizontal sediment mass flux, Q , expressed as the ratio of Q for the ‘clumped’, ‘random’ and ‘street’ configurations relative to Q for the ‘staggered’ configurations at a range of λ and free-stream wind velocities, U_i .

demonstrated that the Raupach et al. (1993) scheme captures the general behaviour of R_t , but to accurately reproduce R_t , both β and m must be tuned for each case. Furthermore, the Raupach et al. (1993) scheme does not accurately reproduce R_t unless β is made variable to suit the roughness configurations (Walter et al., 2012a). This variability is illustrated as the scatter seen in Fig. 3.

Webb et al. (2014) explored the effect of roughness configuration on sediment flux, Q , by comparing Q predicted using the Raupach et al. (1993) scheme using the τ_s PDFs derived by Brown et al. (2008). Webb et al. (2014) found that roughness configuration can have a significant effect on aeolian sediment transport. Surface heterogeneity moderates how much u_* is in excess of u_{*t} (Fig. 4) and therefore both where erosion occurs within a landscape and the magnitude of the total flux from an eroding area. Sediment flux may vary by an order of magnitude for surfaces with the same λ but different roughness configurations (Fig. 5). For very small λ , Q is found to increase with λ , as predicted by Yang and Shao (2005). R_t is found to be sensitive to roughness configuration, and this sensitivity needs to be accounted for in practical applications. The challenges identified by Webb et al. (2014) for implementing the Raupach et al. (1993) scheme for heterogeneous surfaces draw attention to alternative approaches to conceptualising the drag partition that explicitly represent the effect of heterogeneous roughness distributions on wind erosion.

3. Random momentum sinks: From vegetation to saltation

3.1. Owen effect

During saltation, sand grains interact with the airflow and transfer momentum to the surface. The particle momentum flux leads to an increase in roughness length of the aeolian surface similar to the roughness increase induced by waves on the ocean surface (Charnock, 1955) or by the waving canopy of a vegetated surface. This is known as the Owen effect in the aeolian community.

Although the Owen effect was known (Bagnold, 1941), its explanation lacked a solid theoretical underpinning until the work of Raupach (1991). Having worked years on flow over complex terrains, Raupach was naturally very familiar with the studies on vegetation as a momentum sink and immediately recognised that saltating particles behave like stochastic mobile momentum sinks in the saltation layer. For the flow in the saltation layer, saltation reduces the vertical gradient of the flow velocity, and for the flow outside the saltation layer, it increases the capacity of the surface in absorbing momentum thereby increasing z_0 .

By using earlier available observations, Owen (1964) found that the saltation roughness length, z_{0s} , can be expressed as:

$$z_{0s} = A \frac{u_*^2}{2g} \quad (14)$$

with A being approximately 0.02, which is identical to the Charnock (1955) roughness length scheme for 'wavy' surfaces. Eq. (14) is empirical and has two limitations: (a) z_{0s} does not naturally recover z_0 in the case of no saltation; and (b) the observations of Rasmussen et al. (1985) and Gillette et al. (1998) have shown that z_{0s} in the natural environment is much larger than the equation predicts. Raupach (1991) developed an analytical expression for z_{0s} by analysing four inter-related quantities, namely, the mean wind speed, particle-borne momentum flux, air-borne momentum flux and saltation roughness length. Again, to simplify the analysis Raupach made several assumptions in Raupach (1991):

- The total momentum flux is constant in the saltation layer and is composed of a particle-borne momentum flux, τ_p , and an air-borne momentum flux, τ_a , i.e., $\tau = \tau_a(z) + \tau_p(z)$

- τ_p decreases while τ_a increases monotonically with height, and it is required that

$$\begin{aligned} \tau_p(z) &\rightarrow 0 & \text{for } z \rightarrow \infty \\ \tau_a(z) &\rightarrow \rho u_*^2 & \text{for } z \rightarrow \infty \end{aligned}$$

- The characteristic height of τ_p profile, H_s , is on the order of the particle-jump height, such that

$$H_s = b_r \frac{u_*^2}{2g}$$

with b_r being a coefficient; and

- Owen's self-limiting hypothesis for equilibrium saltation applies, i.e., $\tau_a(0) \rightarrow \rho u_*^2$

One functional form for $\tau_a(z)$, which satisfies these constraints is:

$$\left(\frac{\tau_a}{\rho u_*^2} \right)^{1/2} = 1 - (1-r) \exp\left(-\frac{z}{H_s}\right) \quad (15)$$

with:

$$r = \begin{cases} u_{*t}/u_* & u_* \geq u_{*t} \quad \text{saltation case} \\ 1 & u_* < u_{*t} \quad \text{no saltation case} \end{cases}$$

The wind profile in the saltation layer should obey:

$$\frac{\tau_a}{\rho} = K_m \frac{dU_h}{dz} \quad (16)$$

with the eddy diffusivity, K_m , defined as:

$$K_m = \kappa z \sqrt{\frac{\tau_a}{\rho}}$$

Further manipulation gives the wind profile within and above the saltation layer. From the wind profile above the saltation layer, Raupach obtained the Raupach (1991) scheme for saltation roughness length:

$$z_{0s} = \left(A \frac{u_*^2}{2g} \right)^{(1-r)} z_0^r \quad (17)$$

Raupach (1991) suggested that a likely value for A , based on theoretical considerations, is 0.22.

Eq. (17) shows that z_{0s} is a weighted geometric mean of z_0 , the roughness length of the underlying surface and $Au_*^2/2g$. The latter is proportional to the characteristic height of the saltation layer, H_s . In Eq. (17), z_{0s} has two limiting values: when 1) there is no saltation, $r=1$ and $z_{0s}=z_0$, and when 2) there is strong saltation, $r \rightarrow 0$ and $z_{0s}=A H_s$.

At the time when Raupach (1991) was published, little observational data were available to test the scheme. The experiments by Gillette et al. (1997, 1998) at Owens Lake, California provided one of the first tests of the Raupach (1991) scheme. A comparison of the observed and modelled z_{0s} is given in Fig. 6, which shows that the measurements of z_{0s} can be well-described by Eq. (17) using $A=0.38$, a value remarkably close to the predicted value of 0.22. This example is illustrative of many of Raupach's contributions that are built and sustained by a solid theoretical basis but easy to use for the interpretation of observations or the parameterization.

Eq. (15) is the key assumption of the Raupach (1991) model. This assumption is not concerned with how particles move in the saltation layer and has neglected the possible dependence of τ_p on the size of saltation particles. The fact that A is a function of saltation particle size is the likely reason Raupach's first approximation of 0.22 was less than the A value of 0.38 observed by Gillette et al. (1998).

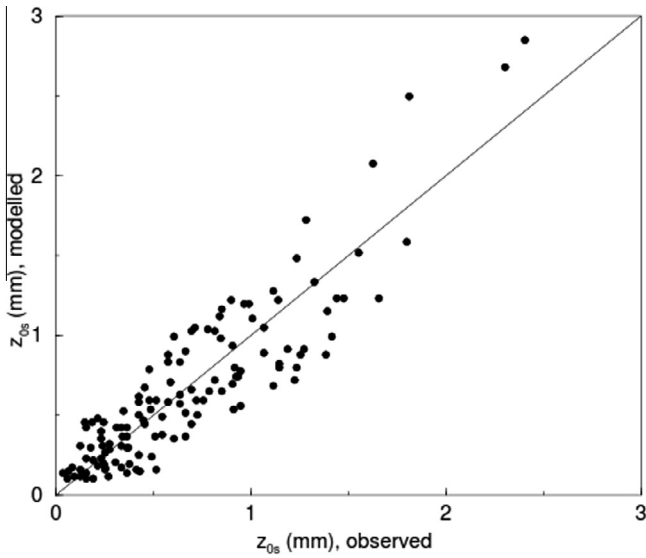


Fig. 6. Modelled saltation roughness length z_{0s} using Eq. (17) versus field measurements of Gillette et al. (1998).

In fact, an infinite number of profiles of τ_a satisfy the requirements proposed by Raupach, but we do not really know how τ_a changes with height in the saltation layer, in that there are very few available direct measurements of τ_a . The measurements of Li and McKenna Neuman (2012) show that shear stress profile in the saltation layer is strongly convex decreasing as the surface of the mobile bed is approached. Another unsolved issue is that Raupach (1991) did not account for the effect of turbulence on saltation trajectories. We can, for example, speculate that increased turbulence should increase the randomness of particle trajectories, and thereby intensify the Owen effect and increase the saltation roughness length. Raupach (1991) may have assumed that the randomness of saltation only causes a secondary effect in particle momentum transfer but this assumption needs testing.

It is not difficult to see that the issues dealt with in Raupach (1991, 1992) and Raupach et al. (1993) are related. In essence, due to the pressure drag on surface roughness elements, roughness length becomes a function of roughness configuration (in the simplest case, frontal area index, λ). In the case of vegetation, roughness elements are plants, and in the case of saltation, roughness elements are randomly moving particles.

3.2. Roughness length, issue of scale, and the albedo analogy

As an extension of Raupach (1992, 1994) proposed a scheme for computing roughness length for climate models. With the simplification of Raupach (1992), Eq. (12) becomes:

$$\kappa^2 \ln^{-2} \left(\frac{z-d}{z_0} \right) = (1 + \beta\lambda) C_s \exp \left(-c \frac{\lambda U_h}{u_*} \right) \quad (18)$$

Eq. (18) is the starting point of the Raupach (1994) scheme, which gained great popularity in the remote sensing community (Schmidt and Dickinson, 2000; Nakai et al., 2008; Tian et al., 2011). Raupach's work inspired Adrian Chappell to make the analogy between aerodynamic sheltering and shadow to retrieve aerodynamic properties over areas to provide a measure that scales linearly from the ground to remote sensing platforms (airborne or satellite) thereby tackling the four challenges of Raupach and Lu (2004) described in Section 2.4.

As discussed in Section 2, the momentum extracted by roughness elements is controlled primarily by λ , i.e., a projection of roughness den-

sity in the direction of wind. When the projection is represented as a zenith angle α , $\tan(\alpha)$ can be seen as a multiplication factor which when restricted to 45° has a value of 1 and results in the projection of the entire frontal area of the roughness element $\lambda_p = \lambda \tan(\alpha)$. To simplify the problem, Raupach (1992) introduced the ideas of sheltering area and sheltering volume that vary with u_* / U_h as shown in Fig. 2 and Eq. (3).

However, it is unlikely that the two dimensional measure λ can adequately characterise the three dimensional nature of aerodynamic roughness (Fig. 7). If geometry projected to the surface is made a function of λ_p with $\alpha = \tan^{-1}(u_* / U_h)$ it should represent the shear stress (u_* / U_h) and the aerodynamic roughness length (z_0 / h). Consequently, Raupach's effective shelter area is changed from a wedge. The new plan-form projection of shadow therefore assumes that u_* / U_h and z_0 / h of wind from a particular direction is dependent on the zenith and azimuth illumination angles. This single scattering albedo was estimated to avoid any dependency on illumination and viewing conditions and to approximate the data available from remote sensing.

Chappell et al. (2010) then showed that the single scattering albedo is related to the z_0 / h from wind velocity profiles of a range of surface roughness conditions in a wind tunnel (Dong et al., 2002). The albedo of the wind tunnel surface roughness was obtained retrospectively by reconstructing a digital elevation model (DEM) of the surface and then ray-casting.

Recent work, with several of the contributors to this paper, has further developed this approach using Marshall's (1971) seminal data. It is now evident that there is a relationship between albedo and many of the essential aerodynamic properties for wind erosion and dust emission modelling (e.g., u_* / U_h). Thus, it appears that this reduced complexity approach inspired by Raupach, enables consistent, repeatable and scalable areal estimates of aerodynamic properties. For example, the global MODIS MCD43A3 albedo product can be used to provide estimates of aerodynamic properties every 500 m and every 8 days between 2000 and present. Fig. 8 shows u_* / U_h and lateral cover (L) for Australia on January 1, 2013.

There appears to be considerable potential for this approach to provide consistent and repeatable estimates of aerodynamic properties and aeolian transport potential. This potential stems from the analogy that albedo and shadow mimic the sheltering effect of roughness and albedo can be retrieved from ground-based (proximal) or remote sensing for large areas, making it a valuable proxy to aerodynamic roughness length. This analogy is probably well justified if roughness elements are sufficiently small compared to boundary-layer depth to exert significant influence on the flow structure, as assumed in Raupach (1992). However, because momentum transfer (governed by the Navier–Stokes equation) and radiation transfer (governed by the radiation transfer equation) have fundamentally different dynamic behaviour (in particular non-linear interactions), it remains to be demonstrated

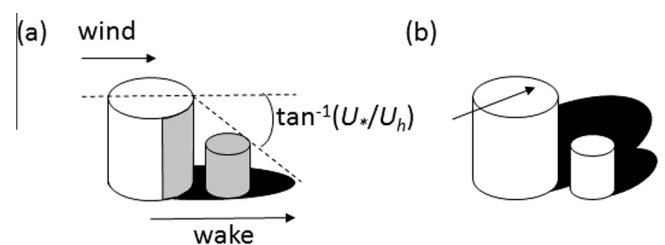


Fig. 7. Roughness protect a portion of the substrate surface (a) that may include all or part of other roughness elements in a heterogeneous surface and following Raupach may be considered dependent on u_* / U_h . A change in wind direction (b) redefines the sheltering effect demonstrating the anisotropic nature of the sheltering.

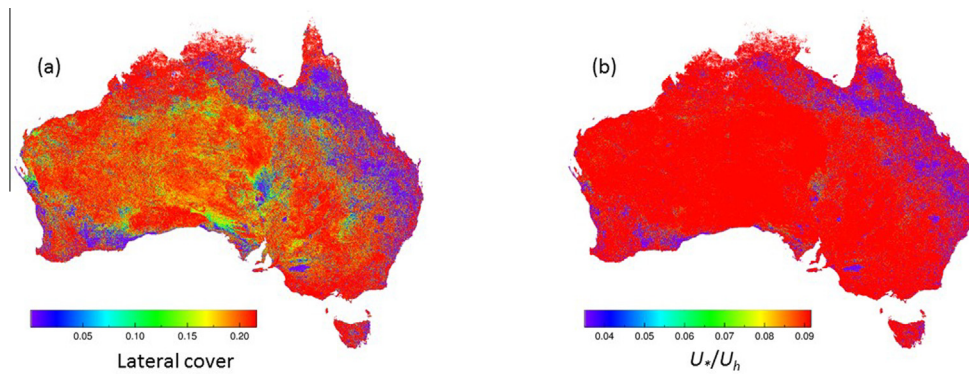


Fig. 8. Examples of aerodynamic properties (a) u_* / U_h and (b) lateral cover estimated from the MODIS MCD43A3 albedo product (500 m resolution) for Australia 1 Jan, 2013.

whether such an analogy exists on a wider spectrum of scales. Nevertheless, it is a prime example of Raupach's inspiration to strive for a practical compromise between parsimony and fidelity in the representation of wind erosion and dust emission modelling of Raupach and Lu (2004) described above in Section 2.4.

4. Dust emission and saltation bombardment

It is already evident in Gillette (1981) that the mechanisms for the entrainment of sand and dust particles differ, because the relative importance of the forces acting on them changes with particle size. The lift-off of sand particles is determined primarily by the balance between the aerodynamic and gravity forces. For smaller particles, the dominance of the gravity force diminishes and the inter-particle cohesion becomes important. It is now known that the gravity force is proportional to d^3 , the aerodynamic force is proportional to d^2 , and although large uncertainties exist in the estimates of cohesive forces, the total cohesive force is proportional to d . For particles with $d < 20 \mu\text{m}$, the cohesive force begins to dominate and hence particles cannot be easily lifted from the surface by aerodynamic forces. Dust particles under natural conditions exist as dust coatings attached to sand grains in sandy soils or as aggregates in clay soils. During weak wind-erosion events, sand particles coated with dust and clay aggregates behave as individuals and the adhering particles may not be released, while during strong wind-erosion events, dust coatings and soil aggregates

may disintegrate, resulting in stronger dust emission. Three dust-emission mechanisms are recognised (Fig. 9):

- **Aerodynamic Lift:** Dust particles can be lifted from the surface directly by aerodynamic forces. As the importance of gravity and aerodynamic forces diminishes for smaller particles and the inter-particle cohesion becomes more important, dust emission arising from direct aerodynamic lift is probably small in general;
- **Saltation Bombardment:** Dust emission is generated by saltation. As saltating particles (sand grains or aggregates) strike the surface, they cause localised impacts that are strong enough to overcome the binding forces acting upon soil dust particles, leading to dust emission. This mechanism is also known as sand blasting or aeolian abrasion (Alfaro et al., 1997; Bullard and White, 2005).
- **Disaggregation:** If saltating grains have dust coatings or if soil aggregates are transported in saltation, the energy exerted on the aggregates during impact can lead to their disaggregation and the release of dust particles. This process is called aggregate disintegration or auto/self-abrasion (e.g., Gillette, 1974; Chappell et al., 2008).

We can formally express the dust-emission rate arising from these three mechanisms as:

$$F = F_a + F_b + F_c \quad (19)$$

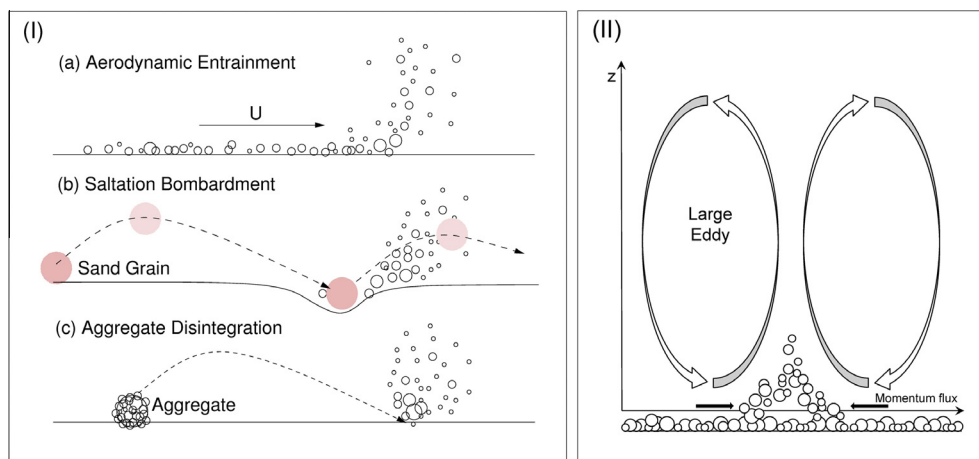


Fig. 9. Mechanisms for dust emission. (I) Dust emission by (a) aerodynamic lift, (b) saltation bombardment and (c) aggregate disintegration. Traditionally, these processes are considered to be driven by mean wind shear, but large eddies can also cause intermittent sand drift and dust emission. (II) Illustration of particle lifting caused by the momentum intermittently transported to the surface by turbulent eddies. Saltation may be involved but does not need to be. (I) modified from Shao (2008) and (II) modified from Klose and Shao (2013).

where F_a denotes aerodynamic lift, F_b saltation bombardment, and F_c aggregate disintegration.

In 1991, Yaping Shao began a postdoctoral position at the Pye Lab under the supervision of Raupach to conduct wind-erosion related research. For conducting the experiments, the portable wind tunnel of John Leys was set up in front of the Pye Lab. Raupach originally planned to test some of the theories that were then developing (e.g., Anderson and Haff, 1991) on saltation feedback (Shao and Raupach, 1992). One day, the then Australian Federal Minister for Environment (Hon. Mr. Ross Free) came to visit the Pye Lab. Raupach et al. were to demonstrate the problem of wind erosion. Soil was placed on the tunnel floor and the wind tunnel was started but no serious dust emission occurred and the Minister was not impressed. The idea of saltation bombardment came to Raupach who then placed sand in front of the dust and produced for the Minister a mini dust storm using the wind tunnel. This story was the origin of the ideas tested in Shao et al. (1993). In that experiment they prepared two beds of material in the wind tunnel: an upstream sand bed which produced a supply of saltating grains, followed immediately by a dust bed that was subject to saltation bombardment. They used combinations of four sand-particle sizes (150, 250, 300 and 600 μm) and three dust-particle sizes (3, 11 and 19 μm). Shao et al. (1993) reported that there was little dust emission even at the maximum flow speed that the tunnel generated ($\sim 20 \text{ m s}^{-1}$) if no saltation particles were introduced, while strong dust emission occurred if sand particles were propelled over the dust surface. Soon thereafter, a similar wind-tunnel experiment was carried out by Alfaro et al. (1997) at the Laboratoire Interuniversitaire des Systèmes Atmosphériques (LISA). Their wind-tunnel experiments not only demonstrated the importance of saltation bombardment on dust emission, but also the emission of more small particles in the case of stronger saltation. What was learned from these experiments is that dust emission is in general proportional to streamwise saltation flux, i.e., $F \propto Q$.

It was soon recognised that the $F \propto Q$ relationship must be soil type dependent. Based on this understanding and using the data of Gillette (1979), Marticorena and Bergametti (1995) proposed the semi-empirical relationship:

$$F = 100 \exp(0.308 \cdot \eta + 13.82)Q \quad (20)$$

where η is percentage of clay content in the parent soil, and F and Q must be, respectively, in $\mu\text{g m}^{-2} \text{ s}^{-1}$ and $\mu\text{g m}^{-1} \text{ s}^{-1}$. Many attempts have been made to develop physically-based dust emission schemes (e.g., Shao et al., 1993, 1996) while it is well-recognised that such efforts are complicated by the fact that the ratio F/Q must also depend on saltation particle size (how much kinetic energy is available) and on soil surface conditions (soft or hard surface, and the strength of cohesive binding forces e.g., Lu and Shao, 1999; Chappell et al., 2008; Kok et al., 2014a).

Attempts soon followed to develop schemes capable of predicting size-resolved dust emission, also called spectral dust emission schemes (e.g., Alfaro and Gomes, 2001; Shao et al., 2011a). The major challenge here is understanding the binding characteristics of dust particles and how they vary in space and time and change with particle size. There is so far insufficient understanding of dust-particle binding strength, but we know from Zimon (1982) that this strength has a stochastic component.

One possible way of overcoming this difficulty is to make use of the observed parent-soil particle size distribution (PSD). It is known from laboratory analysis that minimally dispersed and fully dispersed PSDs $p_m(d)$ and $p_f(d)$ are profoundly different. It is plausible to assume that dust aerosol PSD, $p_s(d)$, is confined by two limits:

$$p_s(d) = \gamma p_m(d) + (1 - \gamma) p_f(d) \quad (21)$$

where γ is the weight for $p_m(d)$ and $(1 - \gamma)$ for $p_f(d)$. Shao (2004) suggested that the emission of dust particles of size d_i arising from the saltation of d_s is given by:

$$\hat{F}_{dt}(d_s) = c_\gamma [(1 - \gamma) + \gamma \sigma_d] (\eta_{ft} \sigma_m + \eta_{ct}) \frac{Qg}{u_*^2} \quad (22)$$

The integration over a range of sand-sized particles gives F_{di} , and the sum of F_{di} over all dust particle size bins gives the total dust emission, F . The process of saltation bombardment is embedded in the parameter $\sigma_m = m_\Omega/m_s$, the ratio between the mass ejected by bombardment, m_Ω , and the mass of the impacting particle, m_s , and in the parameter $\sigma_d = p_m(d_i)/p_f(d_i)$.

Due to the lack of observational data, spectral dust emission schemes were not sufficiently tested earlier. More recently, size-resolved dust fluxes have been estimated from field measurements of dust concentration (Sow et al., 2009). Ishizuka et al. (2014) conducted in Australia a sophisticated field experiment, in which dust emission for several particle sizes was determined. Shao et al. (2011a) were able to use these data to calibrate the Shao (2004) scheme.

Over time a considerable amount of air-borne dust PSD data have been collected around the world. While differences in these PSDs exist, when they are plotted in one graph the differences do not seem to be overwhelming (Fig. 10). This leads to the suggestion, that airborne dust PSD may be universal. There are rational arguments for the approach adopted by Shao (2004), i.e., Eq. (21). However, in hindsight the laboratory measurements of minimally dispersed and fully dispersed PSD do not provide appropriate constraints to $p_s(d)$, because the present-day available $p_m(d)$ is already close to the $p_s(d)$ at maximum saltation intensity, while $p_f(d)$ is simply not achievable through mechanical abrasion.

Although physics based dust emission schemes that require the properties of soil as input for determining size-resolved dust emission is justifiable, this increases the practical difficulty of implementation in large-scale models, such as global climate models

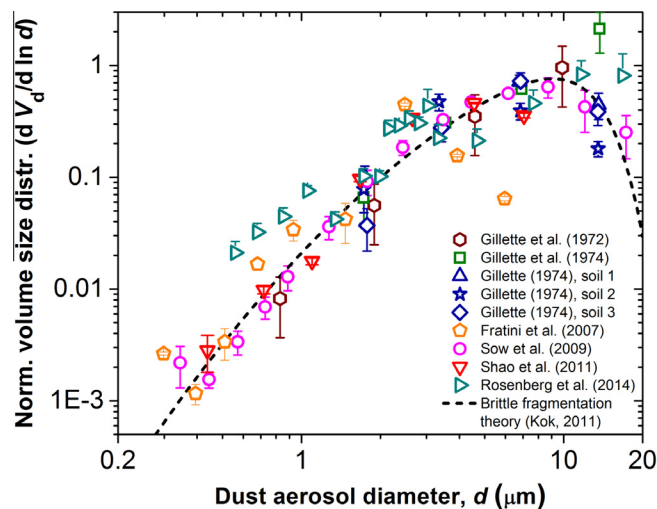


Fig. 10. Compilation of measurements of the volume size distribution of dust aerosols at emission (coloured data), compared with the theoretical prediction from brittle fragmentation theory (dashed line). Measurements by Gillette and colleagues (Gillette et al., 1972, 1974; Gillette, 1974) were taken in Nebraska and Texas and used optical microscopy, whereas measurements by Fratini et al. (2007), Sow et al. (2009), and Shao et al. (2011a) used optical particle counters and were respectively taken in China, Niger, and Australia. All these measurements were made on the ground during wind erosion events. In contrast, the measurements of Rosenberg et al. (2014) were made from an airplane flying over the northwestern Sahara, and used high-frequency optical particle counters to obtain the size-resolved dust flux from eddy covariance. All measurements were normalised following the procedure described in Kok (2011b) and Mahowald et al. (2014).

(GCMs). Consequently, some climate models use ad hoc or empirical assumptions to describe the size distribution of emitted dust aerosols (e.g., Zender et al., 2003; Mahowald et al., 2006; Yue et al., 2010).

Previous research showed that stressed dry soil aggregates fail as brittle materials (Lee and Ingles, 1968; Braunack et al., 1979; Perfect and Kay, 1995; Zobeck et al., 1999). Consequently, Kok (2011b) considered that most dust emission results originated from the fragmentation of aggregates due to saltation bombardment or self-abrasion. Since aggregate fragmentation is a form of brittle fragmentation, the size distribution produced by this process should be scale-invariant for a limited range (Astrom, 2006). The lower limit of this range is set by the size of the aggregate constituent particles, whereas the upper limit is set by the size of the aggregate. Kok (2011b) proposed that the size distribution of dust aerosols can be described by

$$\frac{dV_d}{d \ln d} = \frac{d}{c_v} \left[1 + \operatorname{erf} \left(\frac{\ln(d/\bar{d}_s)}{\sqrt{2} \ln \sigma_s} \right) \right] \exp \left[-\left(\frac{d}{\gamma} \right)^3 \right] \quad (23)$$

where V_d is the normalised volume of dust aerosols with geometric diameter d , c_v is a normalisation constant, σ_s and \bar{d}_s are the geometric standard deviation and median diameter by volume of the log-normal distribution of a typical arid soil size distribution in the 20 μm size range, and the parameter γ denotes the propagation distance of side branches of cracks created in the dust aggregate by a fragmenting impact. Based on measurements of arid soil size distributions (Dalmeida and Schutz, 1983; Goldstein et al., 2005), Kok (2011b) obtained $\sigma_s = 3.0$ and $\bar{d}_s = 3.4 \mu\text{m}$. Furthermore, least-square fitting to dust PSD measurements yielded $\gamma = 12 \pm 1 \mu\text{m}$, such that $c_v = 12.64 \mu\text{m}$.

Eq. (23) is in good agreement with measurements (Fig. 10) of the dust PSD at emission. Note that the newest measurements of Rosenberg et al. (2014) suggest a larger fraction of very fine particles than previous measurements, indicating that more measurements of the dust size distribution are needed. Notably, apart from the Rosenberg et al. (2014) study, the scatter from the different measurements is quite limited, implying that differences in the wind speed and soil size distribution produce only limited variability in the emitted dust size distribution (Reid et al., 2008; Kok, 2011a).

Kok et al. (2014b) developed a new dust emission scheme, the underpinnings of which remains saltation bombardment, now combined with the hypothesis that most dust emission is produced by aggregate fragmentation. Kok et al. (2014b) shows better agreement against a compilation of dust flux measurements than the previous schemes of Gillette and Passi (1988) and Marticorena and Bergametti (1995), both of which are widely used in climate models (Huneeus et al., 2011). Furthermore, the implementation of Kok et al. (2014b) into the Community Earth System Model produces an improved simulation of the dust cycle (Kok et al., 2014a). This improved agreement is at least partially due to accounting for two processes that were not included in previous parameterizations. First, Kok et al. (2014b) accounts for the increasing scaling of dust flux with wind speed that occurs as a soil becomes less erodible and only the most energetic saltators become capable of producing dust. Second, Kok et al. (2014b) accounts for the decrease in dust production per saltator impact that occurs as the soil becomes less erodible. This important effect was previously realised by Shao et al. (1993), and included in the physically-explicit dust emission schemes of Shao et al. (1996), Shao (2001), and Shao (2004), but it is not included in dust emission schemes used in climate models.

The insight that saltator impact speed determines the energy available for dust entrainment is to date an underlying assumption of all dust emission schemes based on saltation bombardment,

irrespective of whether the emission process is then described in terms of energy balance (Shao et al., 1993, 1996; Alfaro and Gomes, 2001), soil dust particle abundance (Marticorena and Bergametti, 1995), volume removal (Shao, 2001, 2004), or fragmentation (Kok, 2011a,b; Kok et al., 2014a,b). The fragmentation process introduced by Kok (2011a,b) gives a specification on the binding energy that scales with particle size, which is consistent with the understanding that inter-particle cohesive force scales with particle size. As we have rather poor understanding of the particle binding strength, the fragmentation assumption offers a reasonable approximation. Given the fact that most observed dust aerosol particle size distributions can be reasonably well represented (Fig. 10), indicates the approximation is useful.

Despite the significant progress made in dust emission modelling during the recent decades, the existing dust schemes contain weaknesses that are still a focus of current research efforts. As Raupach and Lu (2004) already stated in 2004, these weaknesses “include difficulties in application at large spatial and temporal scales, because of input data availability, parameter measurability, and large-scale variability in microphysical parameters and soil properties”.

In most dust emission schemes, u_* and u_{*t} are decisive for the calculation of saltation and dust emission flux. Both u_* and u_{*t} are spatio-temporally integrated quantities and do not describe sub-grid scale and sub-measurement scale variability. No emission is predicted if $u_* < u_{*t}$. However, measurements show that aeolian activities can occur intermittently even if $u_* < u_{*t}$ holds on average (Stout and Zobeck, 1997; Wiggs et al., 2004). Recent studies have focused on intermittent saltation and achieved progress in its numerical modelling. For example, Dupont et al. (2013, 2014) reproduced the development of aeolian streamers due to turbulent eddies by implementing a saltation model in a large-eddy simulation framework.

Aerodynamic dust entrainment has received little attention until recently. This has two reasons: (1) theoretical considerations of inter-particle cohesion suggest that cohesive forces are too strong for particles in the dust-size range to be directly entrained; and (2) dust entrainment without saltation as observed in wind tunnels is much smaller than with saltation. However, considering the stochastic behaviour of inter-particle cohesion due to the multiple influencing factors, such as particle shape, particle surface roughness, or composition, leads to a wide range of scatter even for particles of similar size (Zimon, 1982; Shao, 2008).

A few studies show that dust emission can occur in the absence of saltation, but with much smaller magnitude (e.g., Shao et al., 1993; Loosmore and Hunt, 2000). However, these studies had been set up to evaluate dust entrainment at different mean wind speeds and were not designed to investigate the influence of atmospheric turbulence. Turbulence can have coherent structures induced by buoyancy under unstable atmospheric conditions or by roughness elements as described for vegetation canopies by Raupach et al. (1996). This leads to surface momentum fluxes much larger than the mean wind speed suggests. Convective turbulence is most pronounced in the absence of strong mean winds, i.e., below the saltation threshold. Fig. 11 (Klose and Shao, 2013) shows an example of dust emission generated by convective turbulence modelled with large-eddy simulation. At locations A (micro-convergence lines), B (micro-bursts) and C (vortices), significant dust emission may occur. Due to the stochastic nature of cohesive and turbulent aerodynamic lifting forces, aerodynamic dust entrainment is possible (Klose and Shao, 2012, 2013). In extreme cases (e.g., dust devils), turbulent dust emission can reach the magnitude typical for dust emission induced by saltation bombardment, but in most cases it is typically one to two orders of magnitude smaller. As turbulent dust emission occurs frequently, it may contribute significantly to the global dust cycle (Klose et al., 2014; Li et al., 2014).

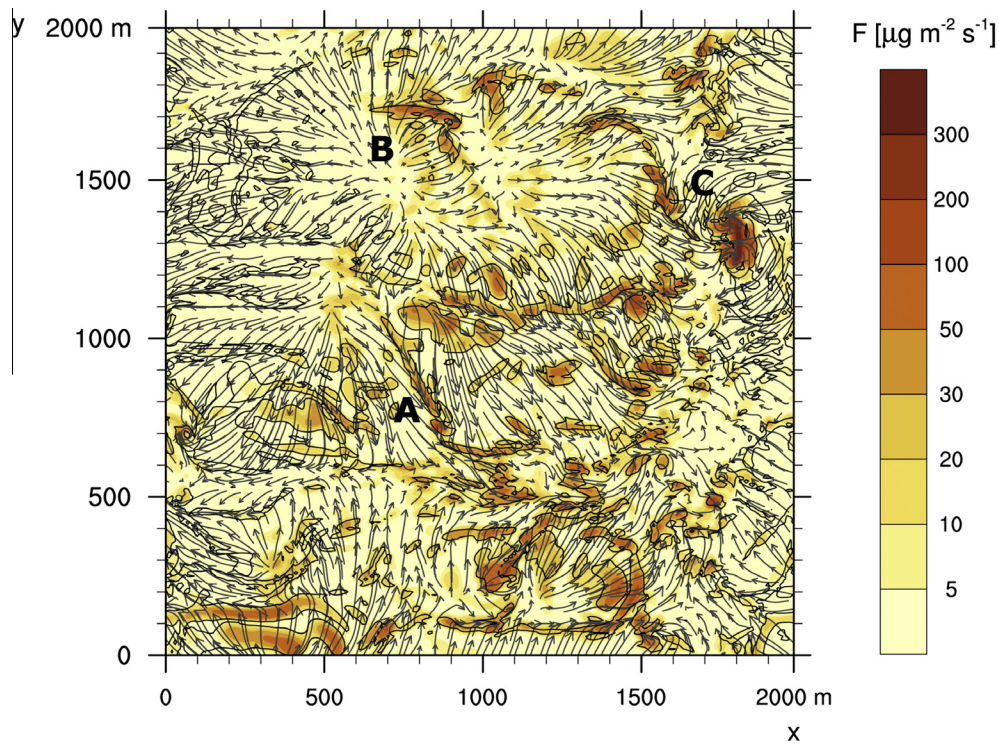


Fig. 11. Turbulent wind speed (vectors, in m s^{-1}) and instantaneous turbulent momentum flux (black contour lines at 1 N m^{-2}) at 10 m height together with turbulent dust emission (shaded, in $\mu\text{g m}^{-2} \text{ s}^{-1}$). Updated from Klose and Shao (2013) by the inclusion of the dust emission scheme of Klose et al. (2014).

5. Threshold friction velocity

5.1. Threshold as control parameter

Shields (1936) studied the threshold friction velocity, u_{*t} , for a spherical particle placed on a bare flat surface, by considering the balance between the gravity force and hydrodynamic drag. He introduced the dimensionless threshold shear stress

$$A = \frac{\tau_{*t}}{(\rho_p - \rho_f)gd} \quad (24)$$

and suggested that A is a function of only the particle Reynolds number, $Re_\tau (= u_\tau d/\nu)$, where ν is kinematic viscosity). In Eq. (24), ρ_p and ρ_f are respectively the particle and fluid density. Bagnold (1941) derived a similar expression for wind-erosion threshold friction velocity and found that for large Re_τ , A is nearly constant and $u_{*t} \propto \sqrt{d}$. Wind-tunnel experiments of windblown sand simulating the atmospheric conditions on Mars and Venus with different kinematic viscosities and/or air densities suggested that Bagnold's expression works well for particles with $d > 100 \mu\text{m}$, but largely under-estimates u_τ for $d < 100 \mu\text{m}$ (Greeley and Iversen, 1985). Iversen and White (1982) pointed out that the rapid increase of u_τ with decreasing particle size is caused by inter-particle cohesion. This led to a revised expression of the dimensionless threshold shear stress A that depends on the inter-particle force, I_p , in addition to Re_τ . The Iversen-White scheme is however rather complex. Shao and Lu (2000) advanced this approach by explicitly considering I_p as inversely proportional to d . This led to a much simpler expression of u_τ with the dimensionless shear stress A remaining as a function of Re_τ only. This new expression has been widely used for estimates of threshold velocity in air and as a reference for other planetary conditions (Burr et al., 2015).

Raupach and Hua Lu collaborated on several research topics, one of which was soil erosion by wind and water. They explored

the question why experimentally derived values of A are consistently higher than the theoretical estimates, and identified several real-world factors that may have major effects on A . These include soil cohesion that can be influenced by temperature and humidity, soil moisture, surface crusting and sheltering effect by roughness elements (McKenna Neuman, 2004). They also considered to what extent temperature-dependent changes in air density and viscosity could play a role in explaining the discrepancies between the observed and theoretical threshold velocities. Such discussion and other topics that related to more general wind erosion modelling led to the review of Raupach and Lu (2004) on the representation of land-surface processes in aeolian transport.

Along this line, Raupach and Lu worked in greater detail to solve the observed puzzle of the Shields' A versus Re_{*t} diagram. When the data obtained from various experiments in air and in water are plotted on the typical Shields' diagram, they do not collapse to a single curve (Fig. 12). What is then the reason for the departure between the data taken in air and water? Lu et al. (2005) proposed a more general expression of A , by incorporating the characteristics of near surface turbulence characterised by the flow Reynolds number $Re_\tau = u_*\delta/\nu$, where δ is the depth of the boundary-layer (Marusic and Kunkel, 2003). They showed that near surface flow velocity increases with Re_τ , and the typical values of Re_τ for air are several orders of magnitude larger than those for water. The large Re_τ in air is associated with intense near-bed turbulence that is dominated by gust-like eddy motions with length scales determined by the characteristic length scale of the roughness (Raupach et al., 1991, 1996). These gusts cause the streamwise velocity to show significant departure from a normal velocity distribution (Morrison et al., 2004), with a strong positive skewness near the bed. Conversely, in water, the mean flow above the layer where the particle entrainment occurs is mostly laminar. This results in a close to normal velocity distribution and smaller length scale of the roughness, therefore smaller values of Re_{*t} and A . They also demonstrated that the upturn of A for small Re_{*t} can also be

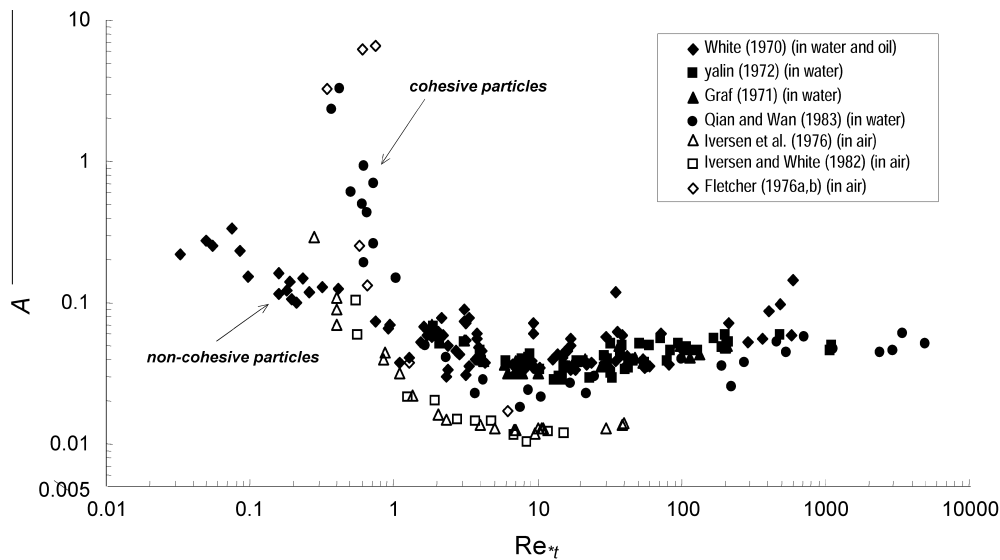


Fig. 12. Dimensionless threshold shear stress A as a function of particle Reynolds number Re_{*t} based on data obtained in water flow (filled) and in air stream (unfilled). These two groups of observations depart both at the large Re_{*t} regime, where aerodynamics dominates and for small Re_{*t} values, where particle cohesion becomes important in determine A . From Lu et al. (2005).

affected by the background flow conditions apart from inter-particle cohesion. As such, they showed that their generalised expression achieves a consistent agreement with data for particle uplift in both air and water flows. Based on their analysis, they suggested that caution is needed in applying previous analytical and semi-empirical models. Perhaps more importantly, they pointed out that incorporating statistical descriptions of the mean flow condition may lead to noticeable improvement of wind erosion models. Indeed, these insights of Raupach and Lu provided the basis for current research based on statistical description, as shown in Klose and Shao (2012, 2013), and some of the aspects considered in Kok et al. (2014b).

6. The carbon link

Soil stores up to 80% of the organic carbon in the terrestrial biosphere and contains more than three times the soil organic carbon (SOC) in the atmosphere (Lal, 2003). The C pools are interconnected and thus a disturbance of the terrestrial C pool (e.g., by soil erosion) can introduce significant changes in the atmospheric C pool. The amount of carbon dioxide (CO_2) captured and converted to SOC annually via terrestrial net primary productivity (NPP) or released as CO_2 by soil microbial respiration (R) is about an order of magnitude greater than the annual increase in atmospheric CO_2 (Houghton et al., 1992). Soil therefore represents a substantial component in the global carbon cycle and small changes in the SOC stock may result in large changes of atmospheric CO_2 (Giorgi, 2006).

Wind-erosion generated dust emission/deposition and the associated SOC exchange between the atmosphere and soil constitutes an important part of the dust-cycle and carbon-cycle interactions, along with the dust-iron effect on the atmosphere and ocean CO_2 exchange (Shao et al., 2011b). For more than two decades (early 1990s to 2015), Raupach worked extensively on the global carbon budget and made a fundamental contribution to that research (Field and Raupach, 2012). Raupach realised early the importance of wind-erosion driven soil nutrient and organic carbon transport, and pointed out that wind erosion removes preferentially the fine, nutrient- and SOC-rich top soil, reduces the soil water holding capacity and thereby causes land degradation (Raupach et al.,

1994). Raupach et al. (1994) provided an assessment of soil nutrient loss, in terms of Nitrogen (N), Phosphorous (P) and Potassium (K), caused by the 1983 Melbourne dust storm. While time did not permit Raupach to work directly on the SOC problem in relation to wind erosion, his initial work was continued by the Australian aeolian research community and in particular Butler, Chappell, Strong and Webb who established the foundation for relating continental estimates of wind erosion to SOC.

For example, Chappell et al. (2014) described how SOC dust emission is omitted from Australian national C accounting and is an underestimated source of CO_2 . They developed a first approximation to SOC enrichment for the dust emission model CEMSYS and quantified SOC dust emission for Australia (5.8 Tg CO_2 -e/y) and Australian agricultural soils (0.4 Tg CO_2 -e/y). These amounts under-estimate CO_2 emissions by approximately 10% for the combined C pools in Australia (based on 2000 estimates), with approximately 5% derived from Australian rangelands and 3% of Australian agricultural soils using the Kyoto accounting method. Northern hemisphere countries with greater dust emission than Australia are also likely to have much larger SOC dust emission. Therefore, omission of SOC dust emission likely represents a considerable underestimate from those nation's C accounts. Chappell et al. (2014) suggested that the omission of SOC dust emission from C cycling and C accounting is a significant global source of uncertainty.

7. Summary and conclusions

In this tribute, we reviewed Raupach's work on aeolian fluid dynamics and the impact of his work on the progress of aeolian research. This is only a small part of Raupach's extensive studies on environmental mechanics and climate change (e.g., Field and Raupach, 2012; Raupach et al., 2014). Specifically for aeolian research, Raupach helped to consolidate the foundation of aeolian fluid dynamics and aeolian modelling, and to propel aeolian research to become a core theme in earth system studies.

Raupach's pioneering work is linked directly to a number of conceptual and modelling advancements made in recent years, while at the same time opening numerous avenues that allowed the aeolian research community to make numerous advances

towards our understanding of aeolian processes. Avenues of inquiry opened by Raupach include:

- (1) *Aeolian Processes over Heterogeneous Surfaces*: Ever since the 1940s, we have focused on studying aeolian processes of relatively simple surfaces, often under the assumption of surface homogeneity and uniform saltation. Thanks to Raupach (1991, 1992) and Raupach et al. (1993), and numerous field and numerical experiments, the essence of the momentum exchange between the atmosphere and aeolian surface is now understood. As we followed Raupach (1992) and Raupach et al. (1993), we realised that the spatial and temporal variations of momentum fluxes profoundly affect aeolian transport, which is in more general terms the typical case of heterogeneous aeolian transport. While research on this topic is rapidly progressing, as demonstrated by Webb et al. (2014) and Dupont et al. (2014), much more needs to be done to establish a theoretical framework and to develop predictive tools.
- (2) *Stochastic and Statistical Dust Modelling*: Existing wind-erosion models are mostly of deterministic nature. However, aeolian processes involve stochastic variables, such as inter-particle cohesion or turbulent surface shear stress, as indicated in Raupach and Lu (2004). New developments in dust-emission models of a statistical nature have been made recently by Klose et al. (2014) and may herald a new generation of wind-erosion models in the coming years.
- (3) *Integration of Aeolian Models with Ecological Models*: The carbon cycle is of central importance to climate studies. Raupach devoted more than 20 years of his academic life to research on the global carbon budget. We now know that understanding of the carbon cycle cannot be completed without knowledge of the dust cycle. This is because dust plays a pivotal role in the atmosphere and ocean CO₂ exchange and aeolian processes are vital for SOC transport and fixation. Thus, aeolian research plays a central role in global Earth system studies. For this to be adequately represented in Earth system models, the dust cycle needs to be better represented (e.g., Kok et al., 2014a,b), but also the coupling of aeolian and ecological processes is important. For steppe landscapes the coupling of wind-erosion models with ecological models is developing (e.g., Shinoda et al., 2011). We expect that this effort will accelerate in the future.
- (4) *New Measurements*: Raupach was famous for his theoretical work, but he was also an accomplished experimental researcher and organizer. He conducted and organised numerous wind-tunnel (e.g., Raupach and Legg, 1983) and field experiments (e.g., Leuning et al., 2004). The very first talk Raupach gave on wind erosion was at the 1st Australian Workshop on Wind Erosion entitled “How to Measure Wind Erosion?”. It was an introductory talk on the basic techniques for saltation measurements. We have moved on since that time, and much more cohesive and sophisticated measurements can be made today. Size-resolved sand transport and dust emission measurements were made early on in Australia (Nickling et al., 1999), in Niger (Sow et al., 2009) and again in Australia in the Japan – Australian Dust Experiment (Ishizuka et al., 2014). New instruments such as PI-SWERI[®] (Etyemezian et al., 2014) and micro wind tunnel (Strong et al., 2015) have been developed for field measurements with emphasis on characterising spatial variability of dust emissions.
- (5) *Large-eddy Aeolian Simulation (LEAS)*: The basic concept of aeolian transport process as a feedback system involving the atmosphere, land surface, and soil particles emerged in

the early 1990s (Anderson and Haff, 1991). Earlier versions of LEAS models were developed by Doorschot and Lehning (2002) and Li and Shao (2003) among others. In more recent years, highly sophisticated LEAS models have been developed, for example, by Klose and Shao (2013) and Dupont et al. (2014). With these models, some of the hypotheses of Raupach can now be fully tested, and more importantly LEAS models serve as powerful tools for generating in depth understanding for improved aeolian process parameterizations (Li et al., 2014; Klose et al., 2014).

For many of us, Raupach was not only a role model scholar, but also a great colleague and a friend. A long-time colleague of Raupach describes that his “excellence in scientific research is not the only skill that enabled Mike to build such a brilliant career. He always had a warm and thoughtful way of collaborating with his colleagues. He showed respect and humility in interacting not only with them, but also with the policy world and the public. Mike’s communications skills were legendary. He could distil the most complex ideas into crisp, understandable stories. His words were carefully chosen, and his spoken sentences often carried the grace and power of expertly crafted written prose. His touchstone, however, was always the science, and in that he was unfailingly rigorous and insightful” (Steffen, 2015).

Raupach was a modest person, always keen to learn from others and at the same time, he was a natural teacher for younger researchers worldwide. He made a large effort to nurture younger Australian aeolian researchers. Harry Butler, Paul Findlater, John Leys, Hua Lu and Yaping Shao all benefited immensely from his deep knowledge and enthusiasm for science. Long after MR² had moved on from aeolian studies, and he was swimming in the much larger research pool of global carbon budgeting, he continued to demonstrate his generosity and nurturing attitude towards students, as for example, when he advised Craig Strong on the fluid dynamics of a micro wind tunnel. In 2014, Strong took up a lectureship at ANU and months later MR² also arrived to take on the role of Director at the Climate Change Institute. Discussion recommenced between them, but sadly MR² passed away before the publication of their work (Strong et al., 2015).

Our community mourns the loss of Raupach as a big thinker and influential leader and as this review demonstrates, his work provided many foundations for the current advances and new directions of aeolian research. As we follow in many of his footsteps and explore uncharted territories, Mike will be missed.

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